# Kinematics of back-arc inversion of the Western Black Sea Basin

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# Kinematics of back-arc inversion of the Western Black Sea Basin

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[1] Back-arc basin evolution is driven by processes active at the main subduction zone typically assuming the transition from an extensional back-arc, during the retreat of a mature slab, to a contractional basin, during high-strain collisional processes. Such a transition is observed in the Black Sea, where the accurate quantification of shortening effects is hampered by the kinematically unclear geometries of Cenozoic inversion. By means of seismic profiles interpretation, quantified deformation features and associated syn-tectonic geometries on the Romanian offshore, this study demonstrates that uplifted areas, observed by exploration studies, form a coherent thick-skinned thrust system with N-ward vergence. Thrusting inverted an existing geometry made up by successive grabens that were inherited from the Cretaceous extensional evolution. The shortening started during late Eocene times and gradually affected all areas of the Western Black Sea Basin during Oligocene and Miocene times, deformation being coherently correlated across its western margin. The mechanism of this generalized inversion is the transmission of stresses during the collision recorded in the Pontides-Balkanides system. Syn-tectonic sedimentation in the Western Black Sea demonstrates that this process was continuous and took place through the onset of gradual shortening migrating northward. Although the total amount of shortening is roughly constant in an E-W direction, individual thrusts have variable offsets, deformation being transferred between structures located at distance across the strike of the system. The Black Sea example demonstrates that the vergence and offset of thrusts can change rapidly along the strike of such a compressional back-arc system. This generates apparently contrasting geometries that accommodate the same orogenic shortening.

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## 1. Introduction

[2] Extensional back-arc basins evolve at convergent plate boundaries in response to the interplay between the convergence of tectonic plates and the velocity of subduction, the roll-back subduction being generally defined as the responsible mechanism [e.g., *Dewey*, 1980; *Morgan et al.*, 2008]. The structural style of back-arc basins can change in time from extensional to compressional with various intermediate classes [*Uyeda and Kanamori*, 1979; *Jarrard*, 1986]. These processes generate sedimentary successions overlying a variety of crustal types, from continental to oceanic ones [e.g., *Molnar and Atwater*, 1978; *Royden et al.*, 1982; *Mathisen and Vondra*, 1983]. Compressional back-arc basins form due to back thrusting of the orogenic core, like in the Pyrenees or Swiss Alps [e.g., *Schmid et al.*, 1996; *Beaumont et al.*, 2000]. Typically, the roll-back driven extension that opens a back-arc basin is enhanced during the mature stage of subduction, due to the gravitational sinking effect of a long and dense slab [e.g., *Doglioni et al.*, 1999]. However, the arrival at the subduction zone of the continental material carried by the downgoing plate during the onset of collision [e.g., *O'Brien*, 2001] changes these general conditions. The buoyant continental crust is gradually involved in shortening, eventually decreasing the roll-back effect, facilitated or not by other processes such as slab break-off, and therefore inverting the earlier formed extensional back-arc [*Horváth et al.*, 2006; *Doglioni et al.*, 2007; *Matenco et al.*, 2010]. Hence, the gradual versus instantaneous transmission of compressional stresses (in the sense of *Ziegler and Cloetingh* [2004]) into large back-arc basins, such as the Pannonian Basin or the Black Sea, locally underlying thinned continental to oceanic crust, is less understood [*Cloetingh et al.*, 2003; *Hyndman et al.*, 2005; *Currie and Hyndman*, 2006; *Horváth et al.*, 2006]. In particular, the balance between the role of the inherited geometry of extensional grabens, subsequent thrust faulting and the formation of a compressional basin is an interesting feature controlling the back-arc evolution during inversion.

[3] Back-arc basins are a common characteristic for all Mediterranean-type of orogens (such as Apennine, Carpathian

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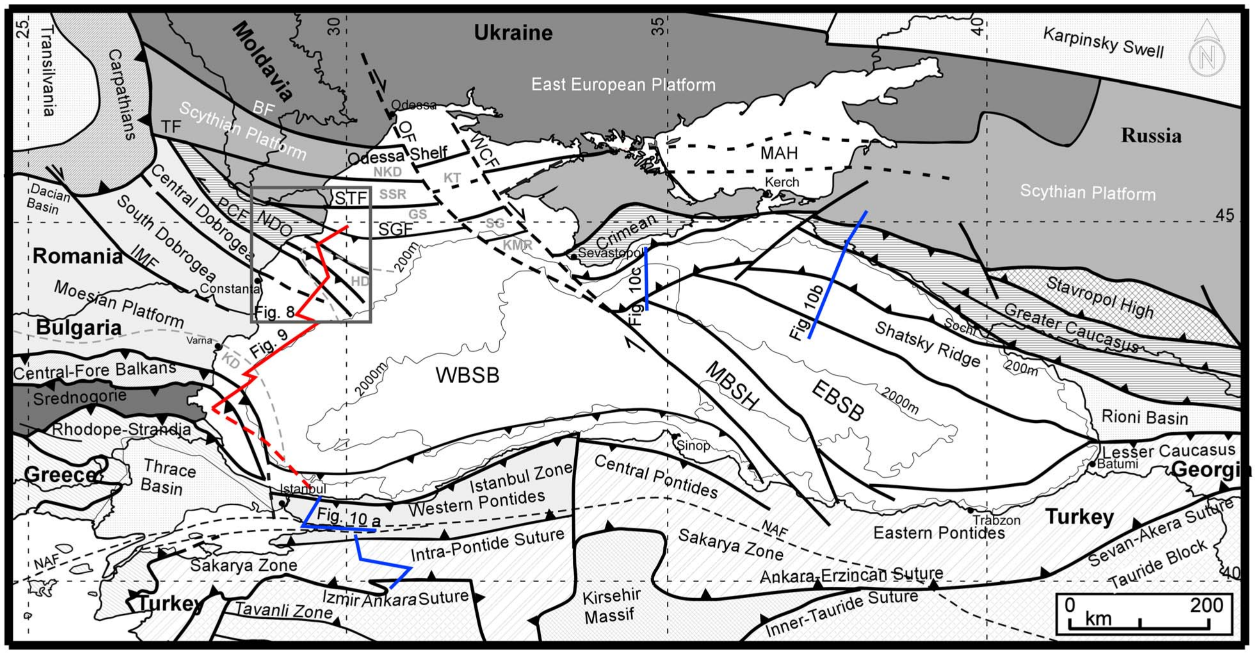



**Figure 1.** Tectonic map of the Black Sea and adjacent areas (compiled from *Finetti et al.* [1988], *Doglioni et al.* [1996], *Robinson and Kerusov* [1997], *Mikhailov et al.* [1999], *Dinu et al.* [2005], *Okay et al.* [2006], *Saintot et al.* [2006], *Afanasenkov et al.* [2007], *Khriachtchevskaia et al.* [2009], and *Nikishin et al.* [2010]). The inset is the location of Figure 8. The red and blue lines represent the location of cross-sections in Figures 9 and 10, respectively. The dotted red line represents the part of cross-section in Figure 9 which is mirrored in order to visualize the same vergence of the Pontide thrust system. Note that the large amount of Paleogene-Neogene faults with small offsets in Crimea, Greater Caucasus and Pontides have been simplified to reflect the most internal and external thrusts and/or back-thrusts of these orogenic belts. The N-ward vergence interpretation of the thrusting observed between the Western Pontides and the Black Sea is taken from *Robinson et al.* [1996]. BF, Bistrița Fault; IMF, Intra Moesian Fault; NAF, North Anatolian Fault; OF, Odessa Fault; PCF, Peceneaga-Camena Fault; SGF, Sfântu Gheorghe Fault; STF, Sulina-Tarhankut Fault; TF, Trotuş Fault; WCF, West Crimea Fault; EBSB, East Black Sea Basin; WBSB, West Black Sea Basin; GS, Gubkin Swell; HD, Histria Depression; KD, Kamchya Depression; KT, Karkinit Trough; KMR, Kalamit Ridge; MAH, Mid Azov High; MBSH, Mid Black Sea High; NDO, North Dobrogea Orogen; NKD, North Kilia Depression; SG, Shtormovaya Graben; SSR, Surov-Snake Island Ridge.

or Hellenide) that are characterized by the coexistence of extension and outward-vergent thrusting at the exterior of highly arcuate mountain chains, as a typical pattern for the Africa/Europe collision zone [e.g., *Faccenna et al.*, 2004; *Brun and Faccenna*, 2008]. The Black Sea is the largest European back-arc basin (Figure 1), its evolution being controlled by processes active during the northward subduction of the Neotethys beneath the Rhodope-Pontides volcanic arc [e.g., *Adamia et al.*, 1977; *Letouzey et al.*, 1977; *Zonenshain and Le Pichon*, 1986; *Okay et al.*, 1994]. It comprises two sub-basins, eastern and western, both being generally through to have oceanic or sub-oceanic crust in the middle part [*Starostenko et al.*, 2004; *Shillington et al.*, 2008] and are separated by the Mid-Black Sea High (Figure 1), which is composed of thinned continental crust [*Stephenson et al.*, 2004]. Following its Cretaceous – early Eocene opening, large-scale compressional episodes are recorded all around its western basin starting with the late middle Eocene collision recorded between the Istanbul and Sakarya blocks [e.g., *Görür*, 1988; *Robinson et al.*, 1996; *Stephenson et al.*, 2004]. So far, a regional integration of the compressional thick-skinned structures observed on the Pontides margin [e.g., *Okay et al.*, 2001; *Sunal and Tüysüz*, 2002], in the Balkanides thrusting [*Doglioni et al.*, 1996], along the Odessa Shelf/Crimean margin [e.g., *Stovba et al.*, 2009] and along the Romanian – northern Bulgarian offshore [*Dinu et al.*, 2005; *Tari et al.*, 2009] is lacking. The latter area displays a significant number of late Eocene – Oligocene structural highs [e.g., *Moroşanu*, 1996], whose unclear genesis has not been yet connected with the coeval inversion observed elsewhere. By interpreting a regional network of (~150) seismic lines acquired for petroleum exploration in the central-western part of the Black Sea basin, correlated with litho-stratigraphic data from (~68) exploration wells, a kinematic and genetic analysis of these structural highs was performed. The analysis focus along the missing link of the Romanian offshore and was complemented with a review of previously published geometries elsewhere, in order to achieve an integrated view along the Western Black Sea basin. This regional integration of inversional structures has been subsequently used to derive the mechanics and geometry of the associated thrust system.

## 2. Constraints on the Opening and Inversion of the Western Black Sea

[4] The Black Sea is situated at the transition zone between a group of orogenic belts formed during the closure of Paleo- and Neo-Tethys oceans and a tectonic mosaic of units deformed in Late Proterozoic to Paleozoic times at the southern margin of the East-European craton [e.g., *Şengör*, 1987; *Okay et al.*, 1996; *Robinson et al.*, 1996; *Stephenson et al.*, 2004; *Saintot et al.*, 2006]. Presently, a number of orogenic systems of various ages are observed along its margins, such as the Istanbul Zone, the Pontides, the Crimea-Caucasus system, the Dobrogea, the Balkanides and Strandja-Sakarya Zone (Figure 1).

[5] The Western Black Sea opened in late Early Cretaceous times (Aptian-Albian) [e.g., *Finetti et al.*, 1988; *Görür*, 1988; *Ţambrea*, 2007], commonly interpreted as a remnant or extensional back-arc basin related to the N-ward subduction of the Neotethys behind the Serbomacedonian – Rhodope – Pontide arc [e.g., *Letouzey et al.*, 1977; *Zonenshain and Le Pichon*, 1986; *Okay et al.*, 1994; *Yilmaz et al.*, 1997; *Afanasenkov et al.*, 2007]. Although a coeval opening of all domains of the Black Sea during Cretaceous times has been inferred [e.g., *Zonenshain and Le Pichon*, 1986; *Nikishin et al.*, 2003], most studies agree that the Eastern Black Sea has opened later, during latest Cretaceous, Paleocene or Eocene times [e.g., *Robinson et al.*, 1996; *Banks*, 1997; *Kaz'min et al.*, 2000] by the rotation of the Shatsky Ridge away from the Mid Black Sea High (Figure 1) [see also *Okay et al.*, 1994], leading to the coeval formation of thinned continental to oceanic crust [*Shillington et al.*, 2008; *Edwards et al.*, 2009].

[6] The Early Cretaceous opening of the Western Black Sea took place in successive deformation phases. This is sedimentologically marked by a gradual change in facies, from Upper Jurassic – lowermost Cretaceous evaporites/carbonates to Lower Cretaceous carbonatic/siliciclastic deposition and, ultimately, by the onset of deep-water sedimentation during Upper Cretaceous times (Figure 2) [*Görür*, 1988; *Dinu et al.*, 2005]. This opening lead to the possible formation of oceanic crust, subsequently followed by a phase of overall subsidence during Late Cretaceous times [*Finetti et al.*, 1988; *Görür*, 1988; *Starostenko et al.*, 2004]. The transition from active rifting to passive margin evolution took place during Turonian – Maastrichtian times, when a generalized transgression and the onset of a widespread carbonatic deposition is recorded, changing from the earlier Cenomanian siliciclastic input (Figure 2). This overall sedimentation pattern was interrupted locally by coarse syn-kinematic siliciclastic deposition (mostly breccias) due to the transcurrent deformation along the Peceneaga-Camena fault system (Figure 1) [*Ţambrea*, 2007]. Hence, the relationship between the successive phases of extensional opening and the transcurrent activity is rather unknown in terms of geometry and kinematic effects. Few differences are observed along the Pontide margin (Figure 1), where the Early Cretaceous back-arc opening was followed by a main extensional episode that took place during Late Cretaceous times, resulted in the formation of large grabens filled with syn-kinematic volcano-clastic sediments [*Görür*, 1988; *Yilmaz et al.*, 1997]. Normal faults can be laterally followed along their strike onshore, where the Late Cretaceous extensional structures are observed in the Srednogorie back-arc basin of the Balkanides (Figure 1). Here, extensional grabens are filled with up to a 3 km thick (volcano-) clastic succession that indicates Campanian deep-water and Maastrichtian – Paleocene shallow-water environments [*Georgiev et al.*, 2001]. Here the extension was associated with the magmatism of the "Banatitic" province, a back-arc calk-alkaline intrusive and extrusive event that extended NW-wards to the Apuseni Mountains of Romania [*Berza et al.*, 1998; *von Quadt et al.*, 2005; *Zimmerman et al.*, 2008]. Near the northern margin of the Western Black Sea (the Odessa Shelf, Figure 1), syn-kinematic deposition in (half-)graben structures demonstrates an Early Cretaceous – early Late Cretaceous age of extension, subsequently followed by a latest Cretaceous – Eocene period of post-rift thermal subsidence [*Khriachtchevskaia et al.*, 2009].

[7] The Eocene opening of the Eastern Black Sea [e.g., *Görür*, 1988; *Meredith and Egan*, 2002] has induced renewed extension in the western basin, which, offshore Romania and Bulgaria, generated faults with offsets in order of tens to hundreds of meters, with associated syn-kinematic deposition [e.g., *Robinson et al.*, 1996; *Dinu et al.*, 2005]. A notable exception is a NE-SW oriented, ~2 km high fault escarpment, located offshore Varna (Figure 1), which is inherited from the initial late Early Cretaceous opening and reactivated during Eocene times, as demonstrated by syn-to post kinematic sediments [*Tari et al.*, 2009]. Hence, the full understanding of this structure origin is still unclear.

[8] Except the Paleocene-early Eocene moments of renewed extension, the overall Upper Cretaceous-Eocene passive margin sedimentation is characterized by a general progradation, the formation of unconformities, the associated system-tracts being driven by sea level variations [*Dinu et al.*, 2005]. This passive margin evolution is ultimately interrupted by the late (middle) Eocene collision recorded southwards between the major tectonic units of the Pontides and the Taurides, smaller continental fragments being accreted in between (Sakarya, Kirsehir [*Şengör and Yilmaz*, 1981; *Okay et al.*, 1994]). This was the time when the last remnants of the Neotethys ocean were closed along the Izmir – Ankara Suture Zone (Figure 1) [*Şengör and Yilmaz*, 1981]. The Pontide - Tauride collision induced large scale uplift that exhumed the southern margin of the Black Sea [*Okay and Sahinturk*, 1997; *Okay et al.*, 2001]. This exhumation provided enhanced siliciclastic influx, making a major change from the earlier dominant carbonatic sedimentation [*Dinu et al.*, 2005].

[9] Existing studies assume a direct prolongation of the Pontides to the Balkanides system onshore Bulgaria along a swing in the thrust system located in the SW part of the Black Sea (Figure 1) [e.g., *Robinson et al.*, 1996]. The Pontides thrusting and exhumation is coeval with the large scale contraction recorded in the Balkanides nappes (the "Illyrian" phase of *Ivanov* [1988]), inverting the Late Cretaceous extensional structures of the Srednogorie and the Eastern Balkans units [*Ivanov*, 1988; *Sinclair et al.*, 1997; *Georgiev et al.*, 2001]. The vergence of the contractional system observed near the contact between the Western Pontides and the adjacent Black Sea (Figure 1) is not yet

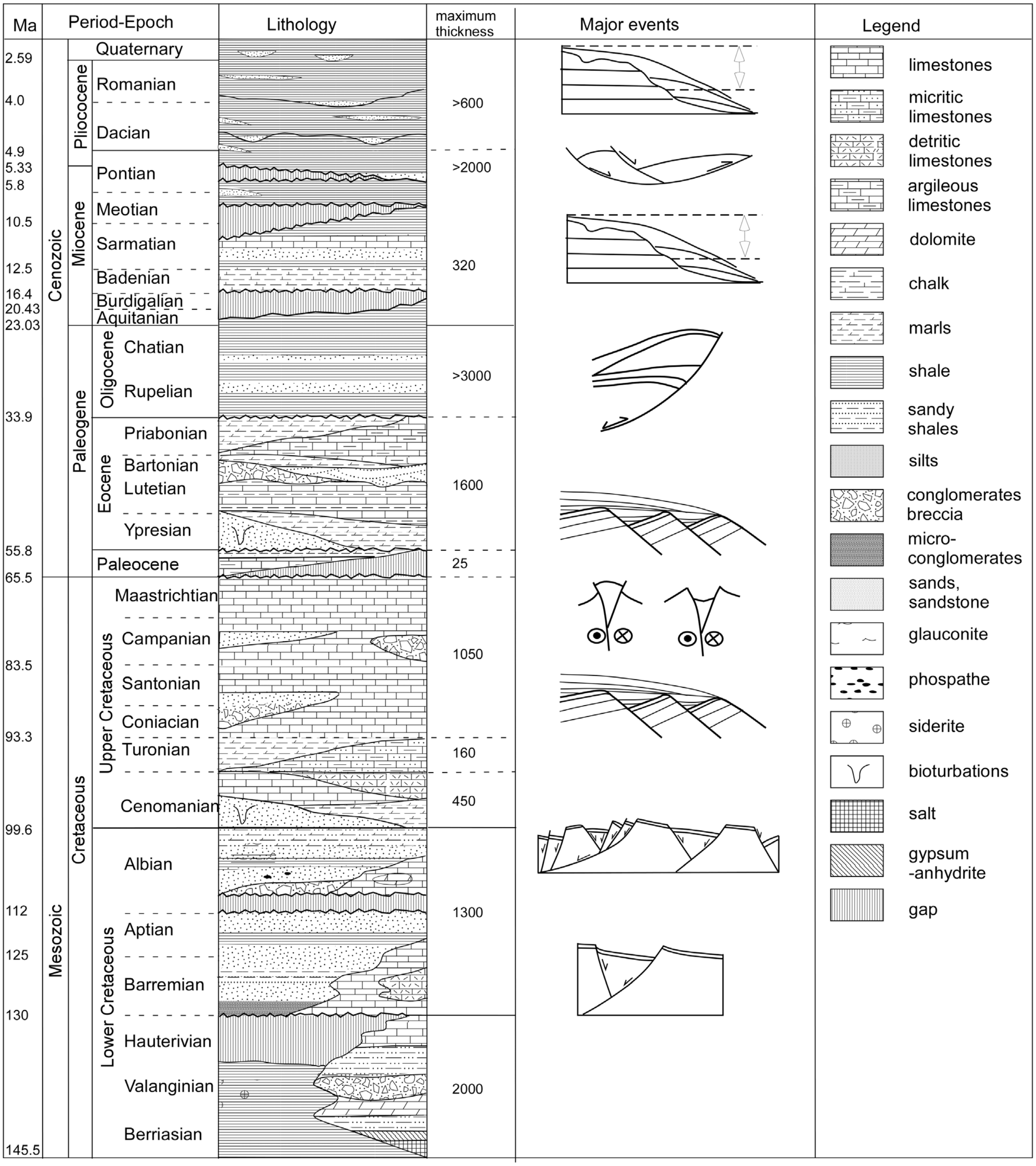
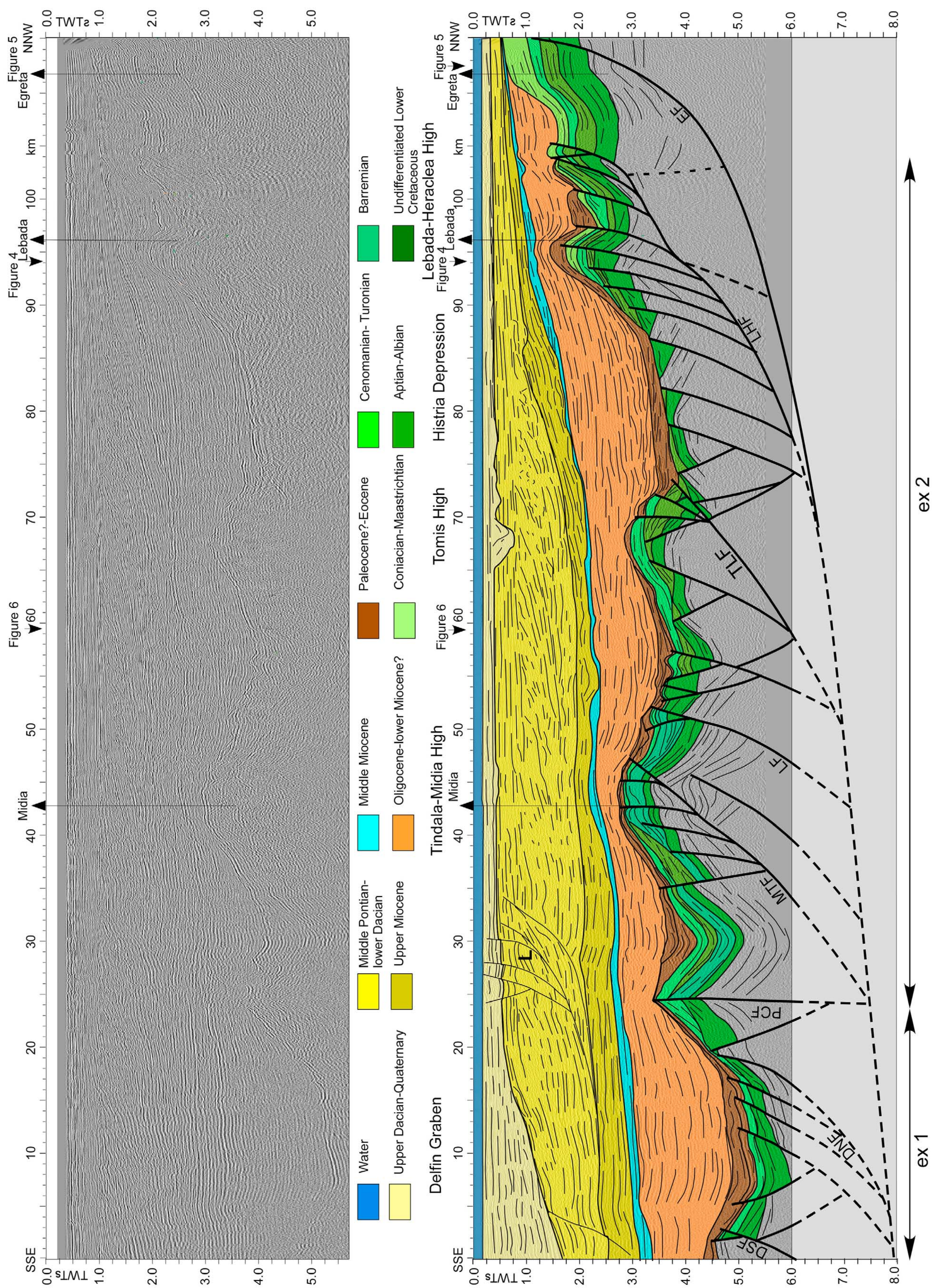



**Figure 2.** Tectono-stratigraphic chart of the western Black Sea part situated in the offshore Romania (compiled from *Băleanu et al.* [1995], *Dinu et al.* [2002], *Ionescu* [2002], *Dinu et al.* [2005], and *Ţambrea* [2007]). The Miocene-Pliocene biostratigraphic ages are a combination between the endemic Central and Eastern Paratethys stages used by the local petroleum exploration (see *Rögl* [1996] for further biostratigraphic correlations). The absolute age equivalents are taken from magnetostratigraphic studies [*Krijgsman et al.*, 2010]. The tectonic events are derived from the present study.

fully constrained. Interpretations generally assume the existence of an association of N- and S-vergent thrusts with reduced offsets [*Finetti et al.*, 1988; *Yilmaz et al.*, 1997; *Sunal and Tüysüz*, 2002]. The Balkanide thrusting was associated in the offshore with a thick late Eocene foredeep that was filled with more than 4 km of sediments in the area of the Kamchya Depression (Figure 1) [*Sinclair et al.*, 1997].



SSE
NNW
TWTs
km
Figure 5
Egreta
Lebada
Figure 4
Figure 6
Midia
Water
Upper Dacian-Quaternary
Middle Pontian-lower Dacian
Upper Miocene
Middle Miocene
Oligocene-lower Miocene?
Paleocene?-Eocene
Coniacian-Maastrichtian
Cenomanian- Turonian
Aptian-Albian
Barremian
Undifferentiated Lower Cretaceous
Delfin Graben
Tindala-Midia High
Tomis High
Histria Depression
Lebada-Heraclea High
EF
LHF
TLF
LF
MTF
PCF
DNF
DSF
L
ex 1
ex 2


Figure 3

[10] The major contraction taking place at the southern margin of the Black Sea led to the onset of inversion recorded in the extensional basins and to the formation of other foreland and thrust-sheet top basins [e.g., *Finetti et al.*, 1988; *Robinson et al.*, 1996; *Sunal and Tüysüz*, 2002; *Nikishin et al.*, 2003]. In contrast with the large-scale structures observed in the south, inversion along the NE part of the Western Black Sea led to the formation of Oligocene-Miocene reverse faults and associated folds with N-ward vergence and offsets in the order of tens to few hundreds of meters [*Afanasenkov et al.*, 2007; *Khriachtchevskaia et al.*, 2009]. Along this northern margin, the structural grain changes rapidly east of the Odessa-West Crimea fault system (Figure 1), where the Crimean Orogen is thrusted S-wards over the Black Sea domain from Oligocene to recent times [e.g., *Stovba et al.*, 2009]. The Odessa-West Crimean fault system has been defined on the basis of potential field geophysics, wells and older seismic reflection profiles to accommodate a dextral displacement of the Odessa Shelf units, such as the Karkinit Trough, the Shtormavaya Graben and the Kalamit Ridge (Figure 1) [e.g., *Garkalenko*, 1970; *Tugolesov et al.*, 1985; *Robinson et al.*, 1996] (or their counterparts in Romanian nomenclature [see *Dinu et al.*, 2005]), when compared with their equivalent units located in Crimea and East Black Sea Basin. More recent interpretations have demonstrated that this dextral structure is characterized in seismic lines by a large zone of deformation made up by high-angle reverse faults with small vertical offsets [*Afanasenkov et al.*, 2007; *Khriachtchevskaia et al.*, 2009].

[11] While the overall Miocene has reduced thicknesses in the entire Western Black Sea, higher rates of subsidence are apparently recorded during the Pliocene-Quaternary times (Figure 2) [*Cloetingh et al.*, 2003; *Nikishin et al.*, 2003]. The well-studied sea level drop of the Messinian Salinity Crisis in the Mediterranean (MSC, 5.96–5.33Ma in the work by *Krijgsman et al.* [1999]) is recorded in the Black Sea by large scale shelf erosion and massive progradation during the subsequent lower Pliocene highstand [e.g., *Hsü and Giovanoli*, 1979; *Gillet et al.*, 2007]. The initial exposure of the shelf margin triggered instability and failure of the outer shelf and slope, generating gravitational gliding and mass transport of sediments toward the deep-sea part of the basin [*Dinu et al.*, 2005; *Ţambrea*, 2007]. Rapid sea level changes are also inferred for the Pliocene-Quaternary endemic evolution of the Western Black Sea [*Winguth et al.*, 2000; *Lericolais et al.*, 2009], when sea level drops triggered the transport of important volumes of sediments toward the deep-sea part of the basin. Consequently, thick successions of mass-transport and turbiditic deposits are observed along a number of deep-sea fans in front of modern rivers discharging into the Black Sea [e.g., *Popescu et al.*, 2001].

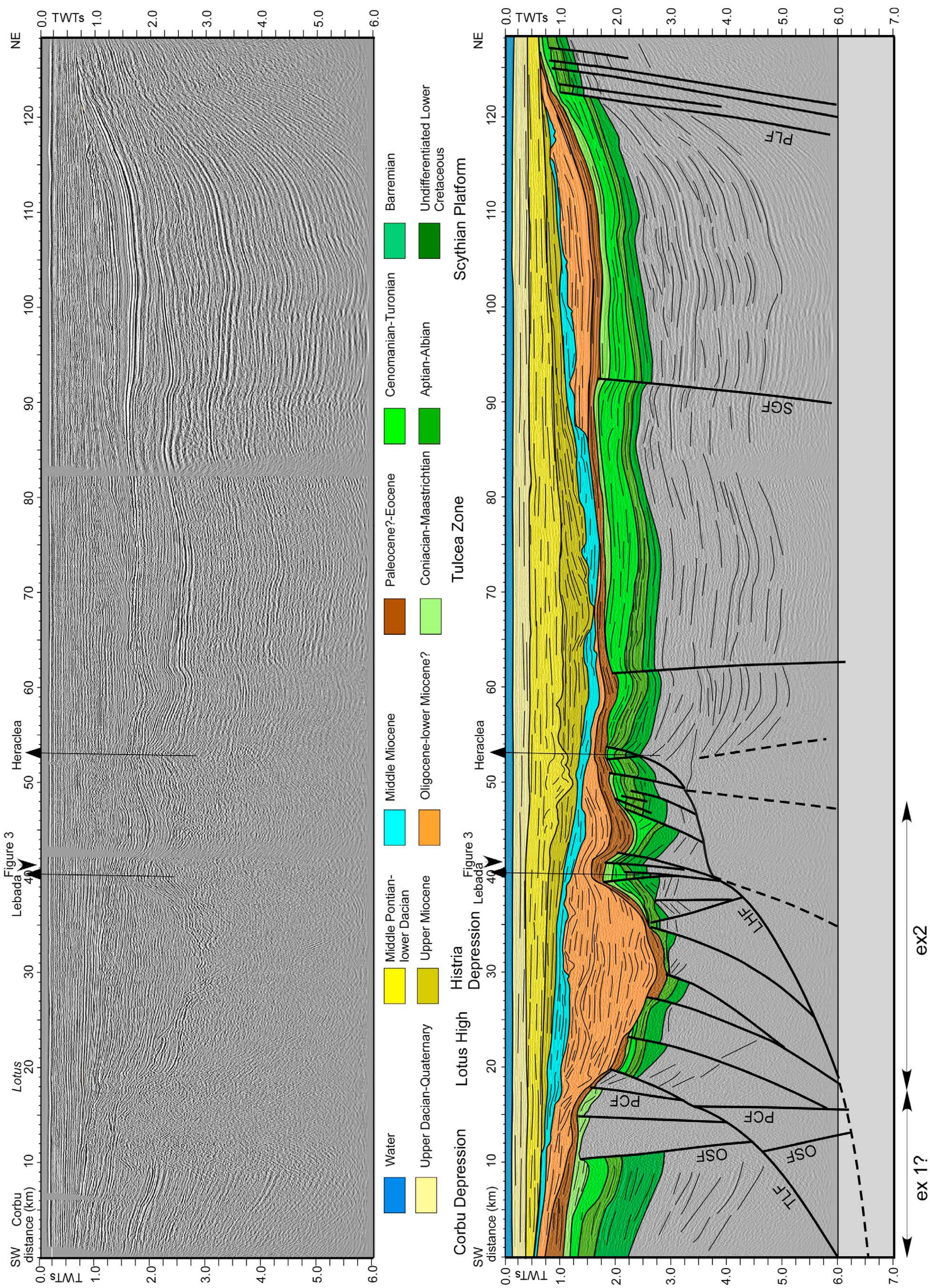
## 3. Structural Interpretation of Seismic Profiles on the Central Part of the Western Black Sea Shelf

[12] The structure of the Western Black Sea passive continental margin indicates complex deformational patterns, buried below younger shelf sediments with almost no deformation. The geometries and syn-kinematic sedimentation demonstrate a Cretaceous period of extension associated with a late-stage of strike-slip movements, and a late Eocene–middle Miocene basin inversion characterized by a significant number of thrusts with relative low offsets (Figures 3, 4, 5, 6, and 7).

### 3.1. Structures Associated With Basin Opening and Enlargement

[13] Normal faults with Barremian-Albian syn-kinematic patterns are coeval with a period of rifting initiation and subsequent basin widening. The geometries are grabens, half-grabens and horsts that were subsequently inverted and are presently oriented WNW-ESE (Figure 8). The observed fault offsets are less than 400 m with wedge-shaped syn-kinematic deposits and local footwall erosion (Figures 3–7). The onset of rifting is observed by seismic syn-kinematic patterns or by facies changes in wells (Figure 2) and appears to be slightly older than the generally assumed Aptian-Cenomanian age [*Görür*, 1988; *Dinu et al.*, 2002]. Wedge-shaped geometries and transparent to chaotic reflectors demonstrate the presence of syn-kinematic sediments, located in the normal faults hanging walls, which formed as a direct result of fault offsets (Figure 3). These first wedges are pre-Aptian in age, as demonstrated by their lateral correlation with wells. Lower Cretaceous facies and thickness changes are frequently related to block asymmetries during normal faulting [*Ţambrea et al.*, 2002]. A facies transition is observed from continental shales during the onset of the Lower Cretaceous to coarse continental and subsequently marine sandstones during Barremian times (Figure 2). These sediments change laterally in terms of lithology and thicknesses: conglomerates and marine sandstones in the normal fault footwalls or areas not affected by normal faulting (like at Corbu, Figures 4 and 7), and a transition from marine sandstones to distal shales located on the normal fault hanging walls (e.g., Midia, Figure 3). These observations support the interpretation of an initial phase of rift initiation (system-tract, in the sense of *Prosser* [1993]) as early as the Berriasian-Barremian times. The grabens are slightly asymmetric, a larger number of faults with higher offsets can be observed along their northern flanks, such as in the case of the Lebăda-Heraclea High (Figures 3 and 4). How-

**Figure 3.** NNW-SSE oriented seismic line and its interpretation calibrated by wells, crossing the shelf of the Black Sea offshore Romania (vertical scale in two-way-travel time, location in Figure 8). Note the series of uplifted structures, thrusts inverting or truncating earlier normal faults, locally organized in half-grabens (Lebăda, Midia). Note also the large thickness of the prograding uppermost Miocene - Pliocene deposits. Ex1 and Ex2 – Cretaceous extensional structures apparently separated by PCF (see section 3.1 for details). CF, Corbu Fault; DSF, Delfin South Fault; DNS, Delfin North Fault; EF, Egreta Fault; LF, Lotus Fault; LHF, Lebăda Heraclea Fault; MTF, Midia-Tîndală Fault; TLF, Tomis-Lotus Fault; OSF, Ostrov-Sinoe Fault; PCF, Peceneaga-Camena Fault; PLF, Pelican Fault; PF, Portiţa Fault; SF, Sinoe Fault; SGF, Sfântu Gheorghe Fault.



SW
NE
distance (km)
TWTs
Corbu
Lotus
Lebada
Figure 3
Heraclea
Water
Upper Dacian-Quaternary
Middle Pontian-lower Dacian
Upper Miocene
Middle Miocene
Oligocene-lower Miocene?
Paleocene?-Eocene
Coniacian-Maastrichtian
Cenomanian-Turonian
Aptian-Albian
Barremian
Undifferentiated Lower Cretaceous
Corbu Depression
Lotus High
Histria Depression
Tulcea Zone
Scythian Platform
TLF
OSF
PCF
LHF
SGF
PLF
ex 1?
ex2


Figure 4

ever, this asymmetric geometry, can also be the result of the subsequent N-ward thrusting, the southern flank of the extensional grabens being largely removed in the thrust hanging walls due to sub-aerial exposure.

[14] Two main extensional structures are separated by the Peceneaga-Camena Fault in the areas of Delfin in the south and spanning from Midia to Heraclea in the north (noted ex1 and ex 2 in Figures 3–7). The first one has the geometry of a (half-)graben filled by syn-kinematic Barremian – Albian deposits in their hanging walls, while the footwalls are locally truncated by subsequent erosion or covered by a post-tectonic cover which starts with the Upper Cretaceous. The second one it is a wider graben-horst structure with tilted normal faults (Figures 3 and 6), which define three smaller grabens separated by two horst structures (Figure 3). The northern flank is formed by a master listric fault dipping to the south associated with numerous synthetic faults (Figures 3 and 4). The southern extension of this structure is the Peceneaga-Camena Fault. The half grabens are filled with syn-kinematic Barremian-Albian and post-kinematic Upper Cretaceous sediments. Syn-kinematic block rotation resulted in exposure and erosion of the uplifted footwalls.

[15] Deformation continued during Late Cretaceous times (Figures 4–6). A reduced number of normal faults locally with larger offsets are observed (such as south of Tomis, Figure 3). Along the flanks of the Heraclea structure, Cenomanian sediments with wedge-shaped geometries located in the fault hanging walls are covered by Coniacian-Maastrichtian sediments (Figures 3 and 4). Late Cretaceous was also the moment when the major Peceneaga-Camena Fault (Figure 1) was (re)activated. This fault is a lithospheric scale structure defined onshore as the southern margin of the Cimmerian Triassic-Jurassic North Dobrogea Orogen (Figure 1), located in the SE European prolongation of the Tornquist–Teisseyre Lineament [*Hippolyte*, 2002, and references therein]. Correlation between individual graben structures located in the vicinity of this major onshore structure and syn-kinematic sedimentation offshore demonstrates a sinistral movement during Late Cretaceous times, although a precise correlation is hampered by subsequent Paleogene truncations (Figures 4, 6 and 7). Strike-slip constraining and releasing bends can be observed along its offshore prolongation (Figure 8), locally with large offsets along individual sub-vertical fault segments. Such a local transpressive structure is the Ostrov-Sinoe Fault, with a ~2 km Late Cretaceous offset that thrusts the metamorphic basement over the thick sediments of the Corbu Depression (Figures 4 and 7). This transpressive movement is associated with high-angle reverse faults that truncate the sediments of the Corbu Depression, which display syn-kinematic depositional geometries during Late Cretaceous times (Figure 7). These faults can be speculated to be the result of a Late Cretaceous phase of compressional deformation. However, the restraining bend geometries observed in map view (Figure 8) have likely resulted from a different mechanism, as being formed by coeval strike-slip deformation recorded along the Peceneaga-Camena and Ostrov-Sinoe faults system.

[16] A number of high-angle normal faults splay-off the Peceneaga-Camena Fault and are coeval with its main strike-slip activity, as demonstrated by the latest Cretaceous syn-kinematic patterns and Paleogene post-tectonic cover (south of PCF, Figures 4, 6 and 7).

[17] One other major fault defined onshore is the steeply dipping Sfântu Gheorghe Fault, along which occurred the thrusting of the North Dobrogea Orogen over the Scythian Platform (Figure 1) [*Seghedi*, 2001]. Offshore, this reverse offset is post-Cretaceous and does not appear to be significant (Figures 4–6).

[18] Successive extensional rifting episodes were gradually replaced by tectonic quiescence during a thermal subsidence stage which lasted until the middle Eocene, locally associated with small-scale differential compaction structures (like as near Lebăda in Figure 3). The overall extensional geometries are truncated or tilted by subsequent late Eocene - middle Miocene thrusts.

[19] A widespread unconformity is observed on top of the Cretaceous deposits, covered by Eocene carbonatic sediments and derived mainly by the interpretation of bio-facies in wells (Figure 2) [*Dinu et al.*, 2005]. This unconformity is often hidden by the deep erosion associated with the onset of thrusting in late Eocene times and its quantification is rather difficult due to the uncertainties related with dating the Paleocene deposits offshore Black Sea (Figure 3) [*Ionescu*, 2002; *Ţambrea*, 2007]. Its origin is unclear, but is most probably related to a combination between shelf exposures during the well-known global sea level drop that took place at the beginning of Paleocene times [*Haq et al.*, 1988] and local uplift recorded along the strike-slip restraining bends that formed in Late Cretaceous.

### 3.2. Inversion Structures in the Western Black Sea

[20] Interpretation of seismic profiles demonstrates that extensional structures and subsequent passive margin sediments in the Western Black Sea were generally inverted starting with the late Eocene times. The inversion resulted in the formation of a large number of widely distributed thrust faults associated with structural highs that separate thrust-sheet top basins. Unconformities observed in wells drilled over the structural culmination of the associated anticlines (Figure 3, around Lebăda, Midia) and the syn-kinematic patterns observed in seismic lines on the flanks of these structural highs indicate that the inversion lasted until middle Miocene times (Figure 5, Egreta and N-wards). Syn-kinematic patterns demonstrate that thrusts offset and growth of the associated anticlines took place gradually: earlier wedge-shaped reflectors were subsequently tilted by later deformation on the flanks of structural highs, while their top was subjected to partial erosion during Oligocene

**Figure 4.** NE-SW oriented seismic line and its interpretation calibrated by wells, crossing the Corbu "depression," Peceneaga-Camena Fault, Tomis-Lotus structure, Histria Depression, Lebăda-Heraclea structure and the Scythian Platform on the Black Sea offshore Romania (vertical scale in two-way-travel time, location in Figure 8). Note the large variations in Oligocene thickness, the overlying erosional morphology and the subsequent truncation of the Peceneaga-Camena Fault. Note also that the low-angle thrust system extends only until the Heraclea structure, the tilted normal faults being only slightly inverted northward.

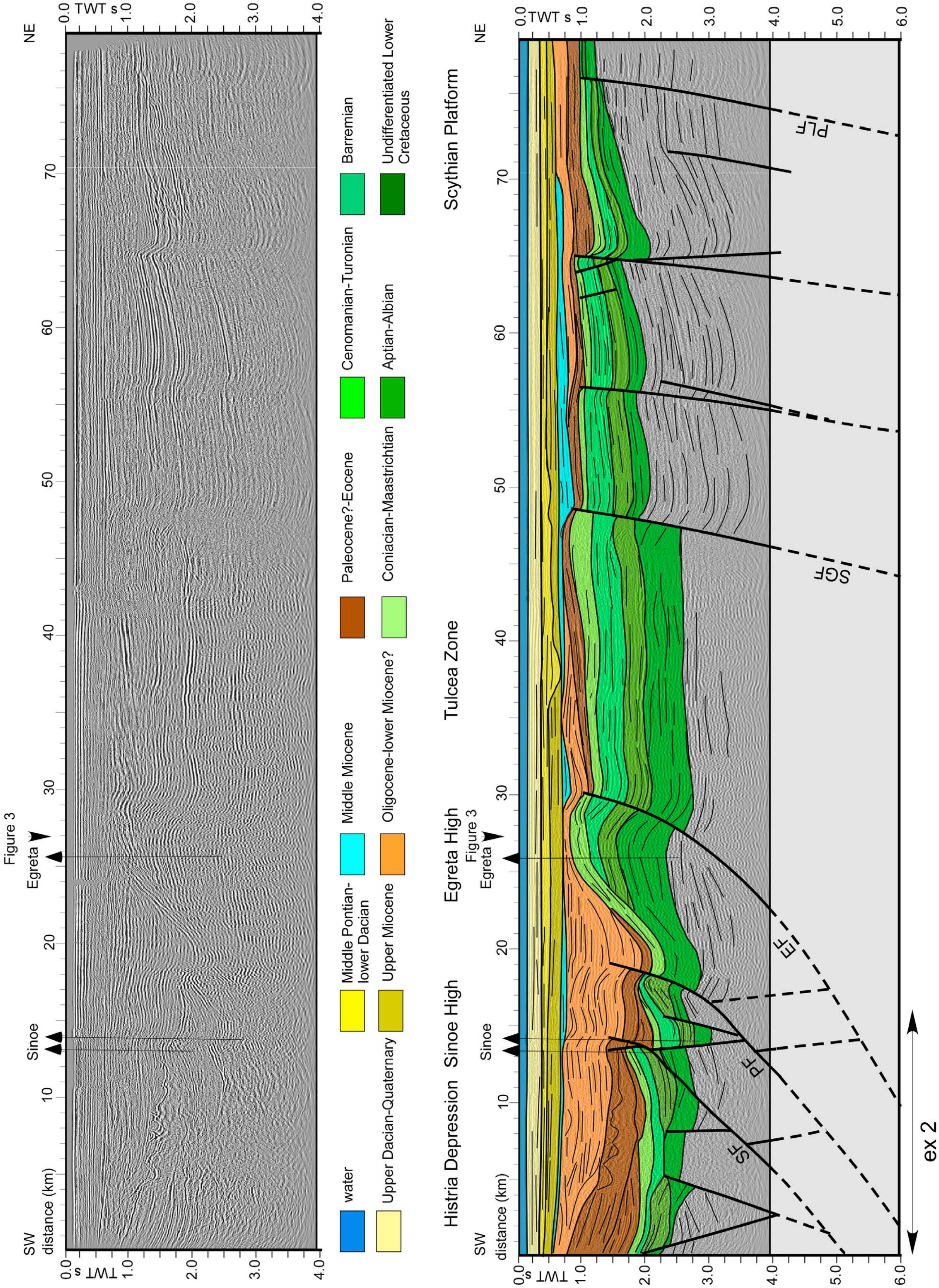

SW
NE
distance (km)
TWT s
Figure 3
Egreta
Sinoe
water
Upper Dacian-Quaternary
Middle Pontian-lower Dacian
Upper Miocene
Middle Miocene
Oligocene-lower Miocene?
Paleocene?-Eocene
Coniacian-Maastrichtian
Cenomanian-Turonian
Aptian-Albian
Barremian
Undifferentiated Lower Cretaceous
Histria Depression
Sinoe High
Egreta High
Tulcea Zone
Scythian Platform
SF
PF
EF
SGF
PLF
ex 2

Figure 5

times (Figure 3 around Lebăda). These syn-kinematic patterns are well preserved, particularly in those places that were subsequently covered by subsidence-driven sedimentation. The inversion formed a large number of N-vergent thrust faults, locally grouped in imbricated fans (Figure 6, north of Delfin). In the latter, reverse offsets formed structural highs along asymmetric, N-verging antiforms. Small S-vergent thrusts (i.e., back-thrusts) are observed along these structural culminations in local pop-up structures (e.g., Figure 3). The thrust system structurally inverted earlier Cretaceous normal faults that dip S-wards, while the ones with N-ward dip were truncated. The sub-vertical position of the truncated normal faults situated in the hanging wall of thrust planes is due to rotation effect coeval with the reverse movement along listric-shaped thrust planes (e.g., Figure 3).

[21] Two late Eocene thrust systems formed due to inversion of the S-ward dipping normal faults of the Cretaceous Delfin Graben (DNS and DNF in Figures 3 and 6), while the N-ward dipping normal faults were most probably truncated by new formed thrust faults. The northward system has a larger cumulated offset in the west as indicated by its pronounced antiformal geometry (Figures 6) when compared with the small offsets in the east (Figure 3). In the former area, syn-kinematic basins can be observed both in the hanging wall and footwall of the structure with wedge-shaped geometries of sediments (Figure 6). The footwall contains tilted reflectors onlapping a lowermost upper Eocene erosional surface, which laterally marks the transition between shallow and deep water carbonates [*Ionescu*, 2002]. At the same place, the continuation of deformation is recorded by yet another erosional unconformity, observed in the middle of the upper Eocene sequence. This later moment of deformation is visible also in the hanging wall basin, where tilted reflectors demonstrate syn-kinematic geometries of late Eocene in age.

[22] The geometries of the thrust lineaments located more to the north indicate that these N-vergent structures were active starting from the late Eocene and continued throughout the early middle Oligocene, as demonstrated by erosion on the top of the antiforms and syn-kinematic patterns along their flank (e.g., Midia, Lebăda, Figures 3 and 4).

[23] The development of Midia-Tîndală thrust system inverted and truncated two different pre-existing (half) grabens of Lower and Upper Cretaceous in age (Figure 3). In this area, the main normal fault controlling the Early Cretaceous half-graben was inverted and eroded on the antiformal culmination with late Eocene syn-kinematic deposition in the thrust sheet top basin carried on the hanging wall flank. Continuation of deformation during the early Oligocene ultimately led to exhumation and erosion of the entire hanging wall. Along strike, the Midia-Tîndală thrust system intersects the oblique Peceneaga-Camena Fault (Figure 6), the thrust hanging wall mostly containing uplifted basement and pre-Cretaceous strata along a wider area of deformation with higher offsets. The Peceneaga-Camena Fault and the Late Cretaceous normal faults on its southern flank are truncated and tilted due to the movement along a listric thrust system (Figure 6).

[24] The thrust lineament of Tomis – Lotus is asymmetric along its strike, from the uplifted and deeply eroded structure of Lotus to the buried antiformal structure of Tomis (Figures 4 and 6, respectively). In the case of Tomis, N-ward vergent thrusts truncated or partly re-activated a graben made up by Early and Late Cretaceous normal faults (Figures 3 and 6). Here, late Eocene and middle Oligocene syn-kinematic sediments (progressively tilted onlaps) observed on the uplift flank are separated by an erosional unconformity, which demonstrates subsidence during thrusting or between the two thrusting events. Among these, the late Eocene deposits in the hanging wall are syn-kinematic as deduced by the observation of clear bi-vergent onlapping reflectors (Figure 3). The Lotus structure is juxtaposed over the Peceneaga Camena - Ostrov Sinoe High (Figure 7). In this case, the N-vergent thrusting truncated an inherited Late Cretaceous strike-slip restraining bend. A back thrusting generated a local pop-up, while progressive tilting of Oligocene reflectors on the northern flank demonstrates continuous syn-kinematic growth (Figure 7).

[25] The most obvious thrust lineament is observed along the Lebăda-Heraclea High, where a large thrust fault has a ramp-flat geometry, which generated a ramp anticline at Lebăda and a fault-propagation fold at Heraclea (Figure 3). The thrusting truncated and/or inverted some of the pre-existing Cretaceous normal faults that extend S-wards to the Tomis-Lotus High (Figures 3, 4 and 6). Other S-dipping normal faults appear to be only carried passively over the ramp anticline at Lebăda (Figure 3). Differences in pre-Eocene strata thicknesses indicate that the main thrust reactivated a pre-existing listric normal fault detach along the northern margin of the Cretaceous graben. Contractional deformation at Heraclea led to reduced offsets along small thrusts that locally form small pop-up structures (Figures 3 and 4). Lebăda and Heraclea structures display syn-kinematic wedge shaped and tilted reflectors in sediments of late Eocene and early – middle Oligocene in age, overlain by post-kinematic late Oligocene strata, similarly to the nearby observed structures. Laterally, deformation along the Lebăda-Heraclea system was transferred by splaying, merging and tear faulting to other, less important thrust structures (Figure 8). More to the offshore, this structure was truncated by subsequent late Miocene normal faults (Figure 6). Toward the onshore, it is recognized in two thrusted and uplifted structures, Sinoe and Egreta (Figure 5). The first one is made up by two individual thrust faults with tilted, wedge-shaped reflectors illustrating a late Eocene and early Oligocene syn-kinematic deposition, separated by a clear unconformity. The second one is a large structure that indicates younger, middle Miocene thrusting, as demonstrated by wedge-shaped reflectors located in late Eocene - middle Miocene sediments and by the unconformity truncating the anticline culmination (Figure 6).

**Figure 5.** NE-SW oriented seismic line and its interpretation calibrated by wells, crossing the Sinoe and Egreta anticlines, and the prolongation of North Dobrogea [see *Seghedi*, 2001] in the offshore (vertical scale in two-way-travel time, location in Figure 8). Note the extension of the low-angle thrust system until the Egreta anticline and the syn-kinematic Oligocene deposits with high variations in thicknesses.

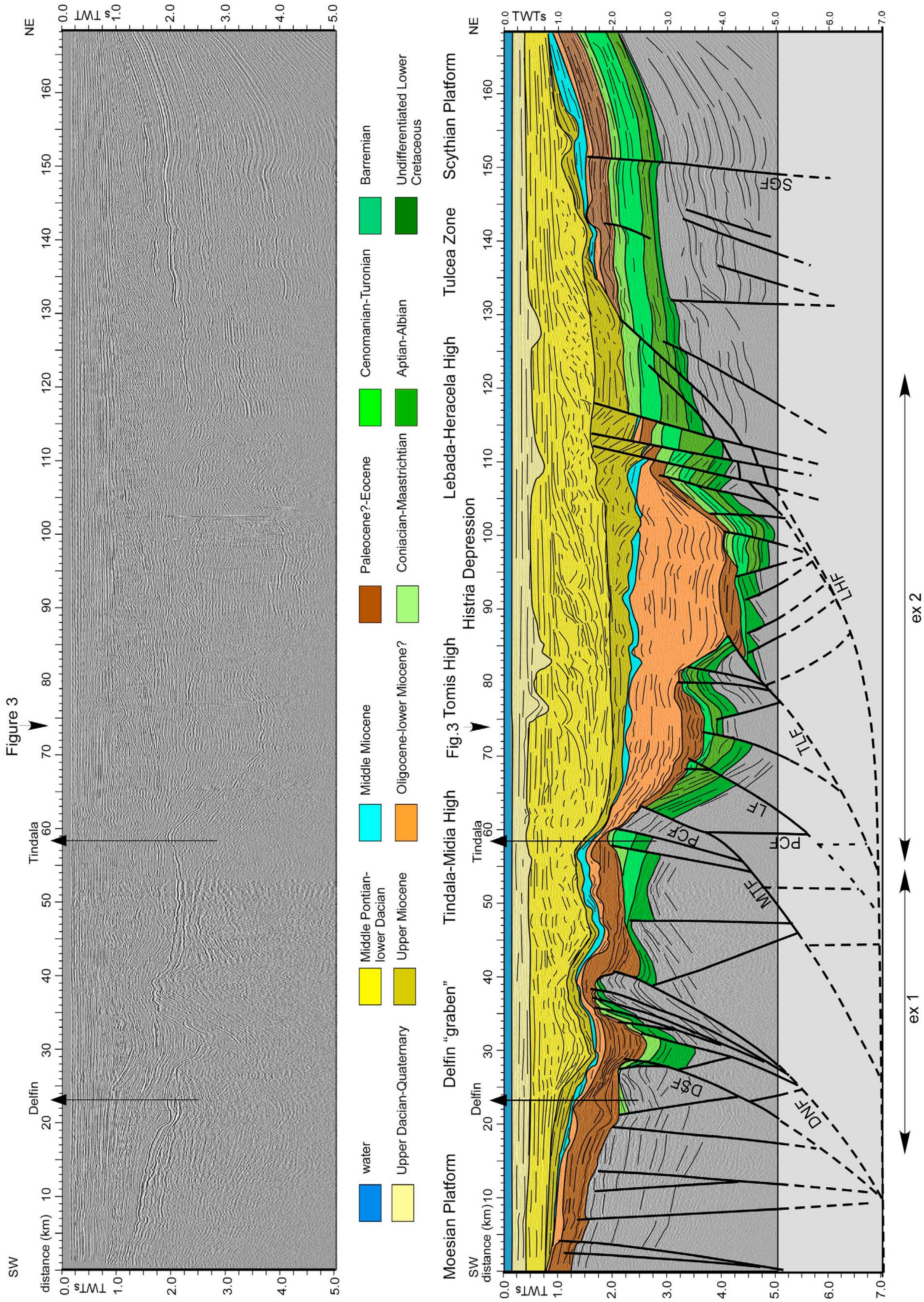

SW
NE
distance (km)
TWTs
s TWT
Figure 3
Delfin
Tindala
water
Upper Dacian-Quaternary
Middle Pontian-lower Dacian
Upper Miocene
Middle Miocene
Oligocene-lower Miocene?
Paleocene?-Eocene
Coniacian-Maastrichtian
Cenomanian-Turonian
Aptian-Albian
Barremian
Undifferentiated Lower Cretaceous
Moesian Platform
Delfin "graben"
Tindala-Midia High
Fig.3 Tomis High
Histria Depression
Lebada-Heracela High
Tulcea Zone
Scythian Platform
DNF
DSF
MTF
PCF
LF
TLF
LHF
SGF
ex 1
ex 2

Figure 6

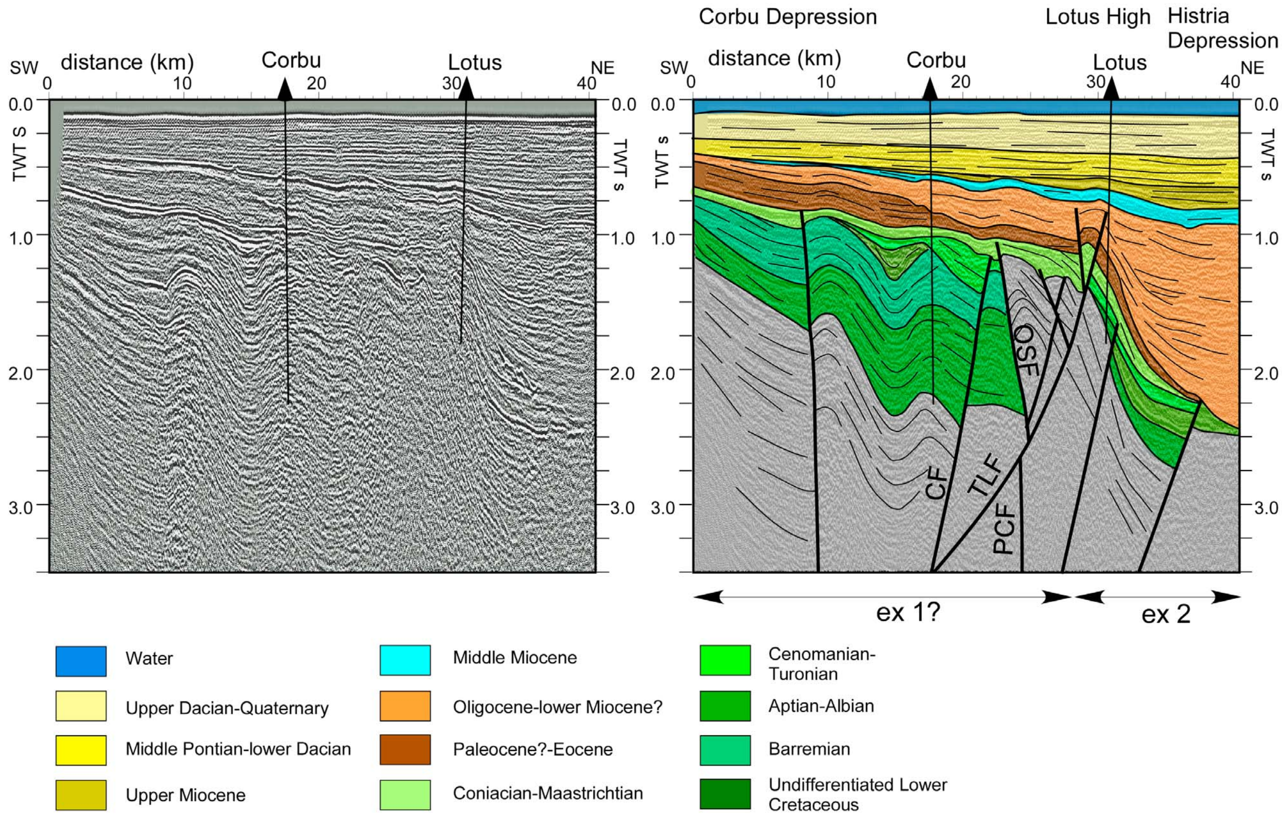



**Figure 7.** NE-SW oriented seismic line crossing the Peceneaga-Camena Fault zone, which separates the Moesian Platform from the North Dobrogea Orogen on the Black Sea offshore of Romania (vertical scale in two-way-travel time, location in Figure 8). An older Cretaceous high-angle fault is truncated by an upper Oligocene thrust, which is associated with a syn-kinematic organization of the reflectors. Note the wedge-type deposition of Cretaceous sediments in the Corbu Depression, which were subsequently folded.

[26] The middle Miocene deformation is increasingly more frequent N-wards (Figures 5 and 6), where small offset faults with N-ward vergence form large open anticlines. Clear syn-kinematic reflector patterns located in Oligocene-middle Miocene sediments and local drag-folding are observed at least on three thrust lineaments (like on the Sfântu Gheorghe Fault, Figures 4 and 5).

[27] The inversion of extensional structures and the formation of new thrust faults during the late Eocene – middle Miocene times are associated with subsidence in thrust sheet top basins. The material eroded from structural culminations is re-deposited in these basins, gradually tilted by ongoing deformation and organized in successive, juxtaposed syn-kinematic wedges. Deposition in these wedges started in upper Eocene (e.g., Delfin and Tomis, Figure 3 and 6), and continued at larger scale during Oligocene – lower Miocene times (Lebăda, Heraclea, Egreta and N-wards, Figure 4). The Oligocene thrusting event is associated with enhanced subsidence as revealed by the deposition of a thick syn- to post-kinematic sedimentary cover of predominantly clastic sediments (Figure 2). The subsidence outpaced the local uplift on top of the emerged anticlines, ultimately burying them beneath this cover. The subsidence is obvious in the case of many thrusted structures, such as Tîndală, Tomis and Lebăda-Heraclea (Figures 3–7), where the total thickness of Oligocene sediments locally reaches 3–4 km. This area with exaggerated Oligocene subsidence spatially corresponds with what was previously defined as the Histria Depression [*Dinu et al.*, 2005]. Elsewhere, reduced amounts of Oligocene subsidence spatially correspond with the zones of low offsets along late Eocene-middle Miocene thrusts (e.g., Figure 5). In other words, a direct spatial relationship is observed between the amount of thrusting and the amount of syn- to post-kinematic subsidence.

**Figure 6.** Regional NE-SW oriented seismic line and its interpretation calibrated by wells, crossing the deformed structures of the Moesian Platform, North Dobrogea and Scythian Platform on the Black Sea offshore of Romania (vertical scale in two-way-travel time, location in Figure 8). Note the relationship between Cretaceous-Tertiary tectonic events, Cretaceous normal faults near Tîndală and north of Tomis structure are inverted/truncated by Paleogene-Miocene thrusts, which are ultimately truncated by late Miocene normal faults. These structures are associated with syn-kinematic deposition, in particular well visible near the Paleogene – Miocene thrusts. Note also the large thickness of the uppermost Miocene - Pliocene deposits, where often channel and/or progradation structures can be interpreted.

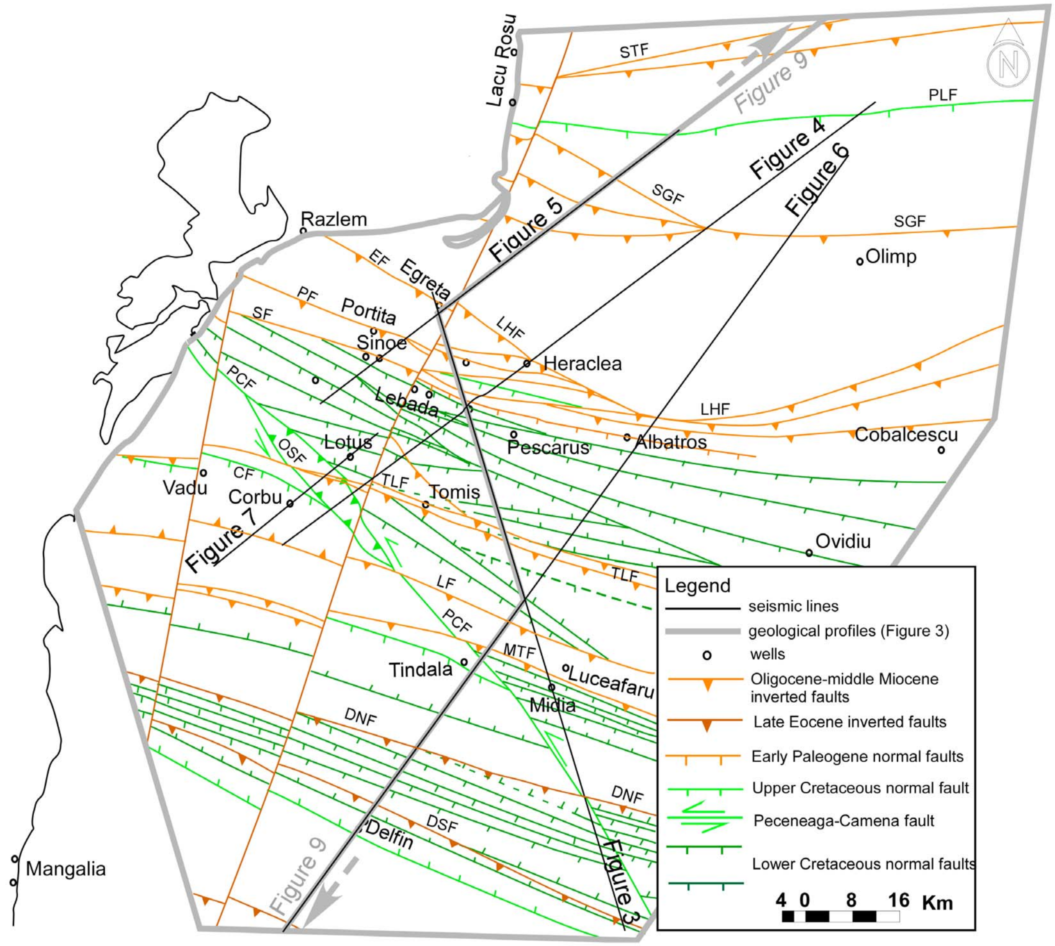



**Figure 8.** Structural map of the Black Sea shelf segment offshore Romania based on the interpretation of regional seismic lines (location in Figure 1). Black lines are the locations of Figures 3–7, gray line is part of transect in Figure 9.

[28] In terms of unconformities, the lower-middle Eocene shallow-water limestones are covered by upper Eocene deeper water siliciclastic-carbonatic sediments (Figure 2). In between, a sub-aerial unconformity is observed by deep erosion, tilting and truncation of reflectors on all seismic lines as the result of thrusting onset during late Eocene times. A large part of the stratigraphic section has been removed, the upper Eocene covering locally directly Lower Cretaceous sediments (such as in Tomis and Lebăda, Figure 3 and 6). Another angular and erosional unconformity is observed between Eocene and Oligocene sediments, locally the erosion removing the entire Upper Cretaceous-Eocene succession (Figure 7). This unconformity is well visible below the thick Oligocene – lower Miocene sediments of the Histria Depression (Figures 3 and 4), enhanced on the flanks and on the anticlines culminations, which demonstrates its tectonic origin. One other marked unconformity is visible at the top of the Oligocene – lower Miocene sequence; overlain by the condensed and aggradational middle Miocene deposits (Figure 4). Biostratigraphic information derived from the interpretation of wells cores demonstrate that the condensed sequence deposition started already during the early Miocene, this stratigraphic interval being generally below the resolution of seismic lines and was largely removed by erosion (Figure 2) [*Dinu et al.*, 2002]. Seismic interpretation suggests that this unconformity was the result of continued contraction-derived uplift superposed on a background of low sediment supply (Figure 5).

### 3.3. Middle Miocene–Quaternary Patterns

[29] The upper parts of the Western Black Sea basin fill displays middle Miocene – Quaternary passive continental margin-type of sedimentation (Figures 3–7). A number of erosional features separating progradational geometries demonstrate the frequent sub-aerial exposures of shelf and continental slope areas that created deep incisions and removed large parts of the older stratigraphic sequence (e.g., Figure 6). Three major unconformities (or sequence boundaries) separate four sequences, i.e., middle Miocene, upper Meotian – lower Pontian, middle Pontian – lower Dacian and upper Dacian – Quaternary (Figures 2, 3 and 6), the associated system-tracts being distinguished on the basis of clinoforms distribution. Wells (like Midia, Figure 3 or Tîndală, Figure 7) demonstrate that the main part of the third sequence (up to 3 km thick; Figures 3 and 6) was deposited in a narrow time interval (middle Pontian – lower Dacian, ~1Ma, see Figure 2). This is in contrast with the couple of hundreds of meters deposited during the pre-dating ~9Ma

middle Miocene first sequence and the < 1 km thick sediments deposited during the post-dating ~4.5Ma upper Dacian – Quaternary fourth sequence. This geometry is interpreted to be the result of sea level variations and their relationship with a basin situated between the Black Sea and the Carpatho-Balkanic source area, i.e., the Dacian Basin (Figure 1) (I. Munteanu et al., Quantifying the effects of large sea-level variations in connected basins: The Dacian–Black Sea system, submitted to *Basin Research*, 2011). As long as an accommodation space for sediments was available in the Dacian Basin, it functioned as a trap for the sedimentary flux generated by the source area. When the Dacian Basin was completely filled during the large scale Messinian Salinity Crisis (MSC) sea level drop, which took place in the Eastern Paratethys during the equivalent intra-Pontian times, large amounts of sediments were discharged into the Black Sea with typical progradational geometries (Figure 6). During periods of low-stands, the deep shelf incision observed along large canyons is associated with the formation of a thick succession of mass-transport and turbiditic deposits (in the sense of *Posamentier and Walker* [2006]) recorded along a number of deep-sea fans in front of modern rivers, like the Danube or Dnieper [*Popescu et al.*, 2001].

[30] In more detail, a larger amount of second-order sedimentological features are visible in seismic lines such as numerous erosional surfaces, channels, gravitational landslides and higher order sequence boundaries (Figures 3–5 and 6). The gravitational landslides are visible by décollement zones and chaotic facies (L, Figure 3) and are interpreted as slope instabilities due to sub-aerial exposure of shelf and continental slope areas during the MSC event. Movement along décollement zones created tilted reflector geometries covered by progradational ones during the subsequent highstands. A small number of normal faults, of apparent tectonic origin, truncate the earlier thrust system, are associated with syn-kinematic sedimentation and are sealed by the intra-Pontian unconformity (Figure 6). The uppermost unconformity shows spectacular erosional features along shelf canyons feeding a low-stand deep-sea fan-type of sedimentation, which were filled during the subsequent transgressive event (Figure 6). A careful analysis of the biostratigraphic information in the available offshore wells demonstrates that the unconformity marks another slightly younger sea level drop event of intra-Dacian in age (Munteanu et al., submitted manuscript, 2011).

## 4. Inversion of the Western Black Sea Back-Arc Basin

[31] A number of studies describes late Eocene – middle Miocene inversional structures that have been observed in various other locations along the Western Black Sea margin [*Doglioni et al.*, 1996; *Moroşanu*, 2002; *Stovba et al.*, 2009]. However, the Romanian sector has not been so far integrated in a well-defined framework of thrusting geometries and, therefore, a correlation of basin inversion between the Pontides-Balkanides and the northern Odessa Shelf is required in order to understand the system kinematics.

### 4.1. Timing of Inversion

[32] The studied Romanian offshore demonstrates a gradual onset of thrusting during the late Eocene - middle Miocene, deformation migrating in time northward in what appears to be a foreland-propagating sequence in respect to the Pontides. Along this first order pattern, some individual thrusts were locally activated out-of sequence, like the Tîndală High (Figures 6 and 9).

[33] The positive inversion started during late Eocene times, the associated thrusts displaying reduced offsets, as observed by the syn-kinematic geometries in the area of Delfin and Midia highs (e.g., Figure 6). This was coeval with the main inversion episode of northward thrusting recorded in the Balkanides and their offshore prolongation. Further southward, this deformation was coeval with the thrusting recorded in the frontal part of the Istanbul Zone, offshore Turkey (Figures 1 and 9) [*Finetti et al.*, 1988; *Doglioni et al.*, 1996; *Georgiev et al.*, 2001]. More to the southern onshore, the Istanbul Zone and Sakarya Zone display the large scale contractional deformation associated with the closure of the Izmir-Ankara Suture (Figure 10a) [*Yilmaz et al.*, 1997; *Görür and Tüysüz*, 2001]. The bulk of the deformation here is considered late Eocene, although post-tectonic covers lack in many locations and a later post-Eocene shortening-related uplift cannot be excluded (Figure 10a) [*Sunal and Tüysüz*, 2002].

[34] The main bulk of deformation took place in the studied area during Oligocene – early Miocene times when uplifted areas formed in or near the Histria Depression (Figure 9). No significant coeval contractional deformation is recorded in the onshore part of the Balkanides, as demonstrated by the widespread observation of a post-tectonic cover [*Ivanov*, 1988], while their offshore prolongation display tilted Oligocene geometries, suggesting limited offsets (Figure 9) [*Doglioni et al.*, 1996]. Northward, the Oligocene – middle Miocene times correspond with an early onset of thrusting on the Odessa Shelf [*Khriachtchevskaia et al.*, 2010], where similar structural geometries (high angle reverse faults) are observed as in the northern part of the studied area (Figure 9).

[35] The post - middle Miocene contractional deformation is fairly reduced and recorded only on the Odessa Shelf, at the northernmost extremity of the Western Black Sea (north of the section in Figure 9). Wide open anticlines or smaller-scale folds formed along the same type of high-angle reverse faults with reduced offsets activated earlier during the Oligocene – early Miocene [*Afanasenkov et al.*, 2007; *Khriachtchevskaia et al.*, 2010].

[36] The overall northward migration of thrusting along the Western Black Sea shelf is reflected also by the formation of associated foredeeps (Figure 9). While the bulk of foredeep-wedge geometries in the frontal part of the Pontides, the Balkanides (Kamchya Foredeep) and southern Romanian offshore (Delfin area) is late Eocene in age, these types of geometries migrate in age to Oligocene – lower Miocene in the area of the Histria Depression and furthermore to middle Miocene in front of the small thrusts observed northward (Figure 9).

### 4.2. Lateral Variation in Shortening Along the Western Black Sea

[37] The thrusts observed in the studied area are trending WNW-wards toward the onshore. However, the late Eocene – middle Miocene thrusting cannot be correlated with similar structures potentially located onshore. The only exceptions to this rule are the Peceneaga-Camena and Sfântu

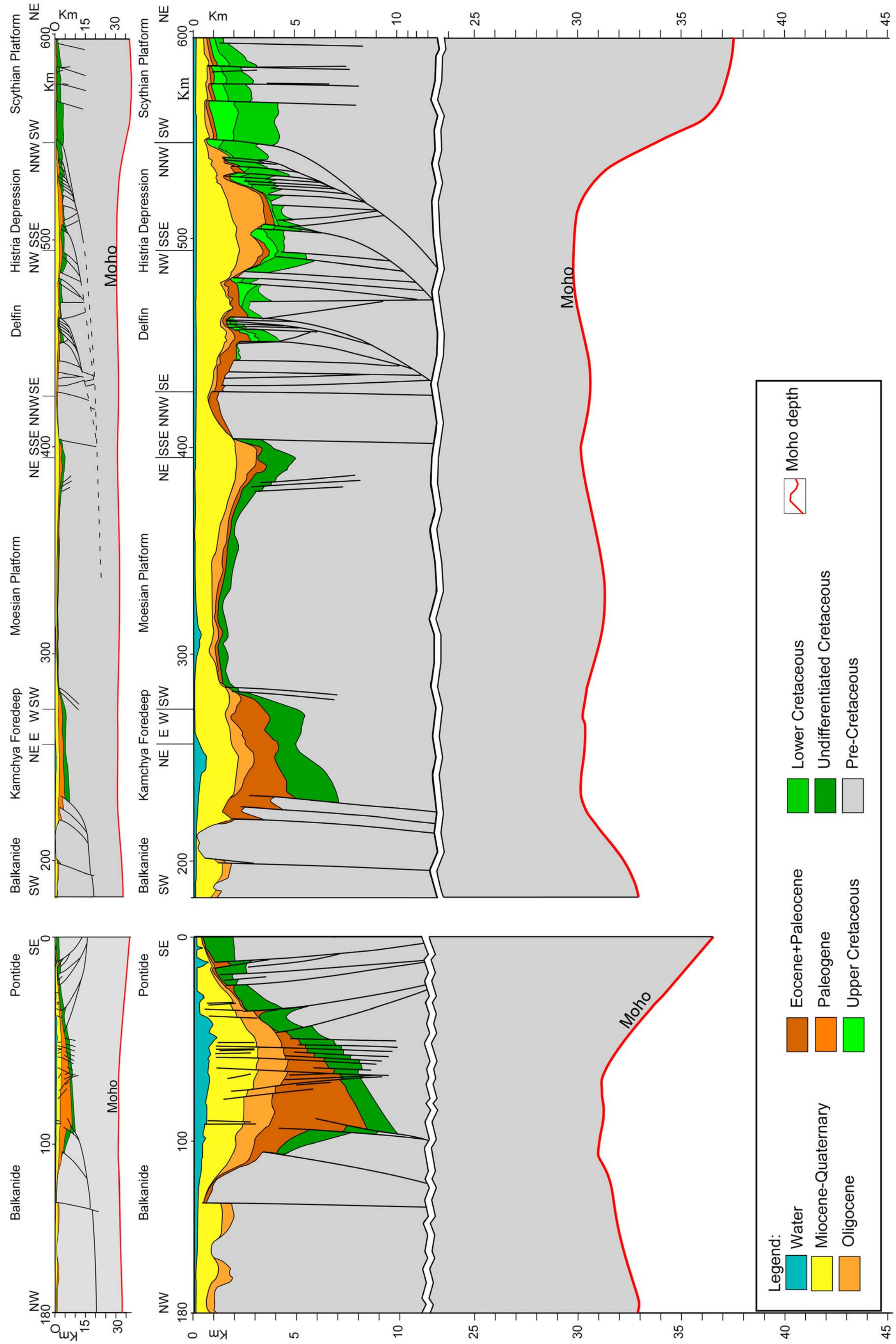

NW
SE
Balkanide
Pontide
Moho
SW
NE
Kamchya Foredeep
Moesian Platform
Delfin
Histria Depression
Scythian Platform
NE E W SW
SSE NNW SE
NW SSE
NNW SW
Km
Legend:
Water
Miocene-Quaternary
Oligocene
Eocene+Paleocene
Paleogene
Upper Cretaceous
Lower Cretaceous
Undifferentiated Cretaceous
Pre-Cretaceous
Moho depth

**Figure 9**

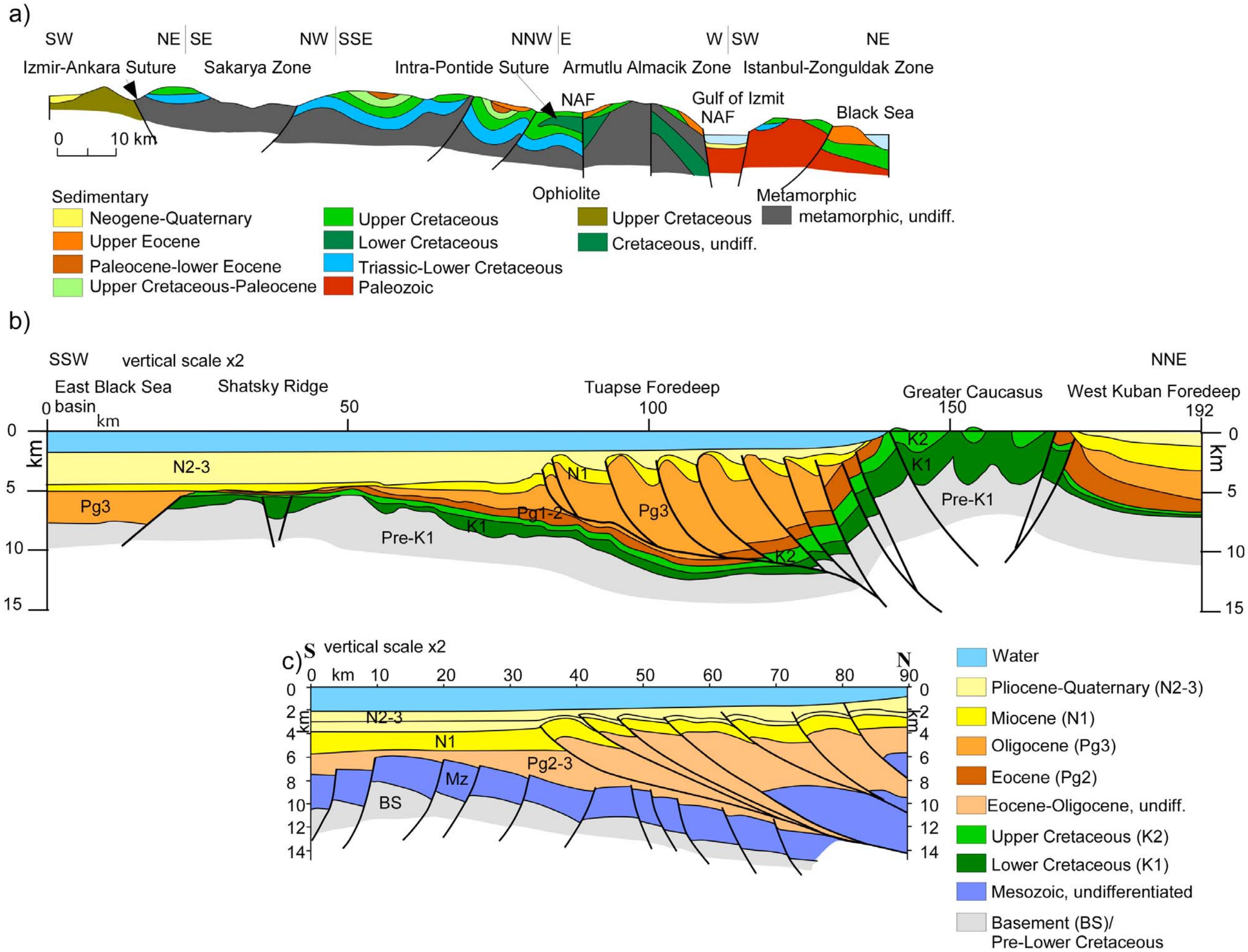



**Figure 10.** (a) Regional section connecting the structures offshore Romania displayed in Figure 8 with the onshore kinematics of the Pontides [see *Yilmaz et al.*, 1997, Figure 1]. (b) Regional cross section over the Greater Caucasus and the NE part of the Eastern Black Sea [see *Nikishin et al.*, 2010, Figure 81]. (c) Interpreted seismic line converted in depth showing the thrusting of Crimea over the Eastern Black Sea Basin [see *Finetti et al.*, 1988, Figure 6]. This structure was confirmed by similar, modern seismic interpretations offshore Crimea [*Afanasenkov et al.*, 2007; *Stovba et al.*, 2009].

Gheorghe faults (Figures 5 and 7), and a couple of reverse faults with minor offsets described in South Dobrogea [*Hippolyte*, 2002; *Dinu et al.*, 2005]. On the overall, the total amount of shortening recorded onshore is in the order of 1–2 km. This is in contrast with the restoration of thrust offsets and associated folds (Figure 9) observed of the Tîndală-Midia, Tomis-Lotus and Lebăda-Heraclea highs, which indicate 3–4 km individual shortening values and cumulate a total amount of 15 km (central zone in Figure 11). The gradual reduction of shortening toward the onshore is observed by gradually reduced offsets along the same thrusts in the areas of Sinoe-Portița highs and Corbu Depression (Figures 7 and 8), cumulating a total shortening in the order of 6 km (western zone in Figure 11). Therefore, the shortening gradually dies out toward the WNW-ward located onshore and decreases also ESE-ward. The latter is demonstrated by individual structures in the areas of Albatros and WSW prolongation of Delfin Graben (Figure 8) that have

**Figure 9.** (top and bottom) Regional transect over the western part of the Black Sea starting from the Istanbul Zone (offshore western Turkey) to the Scythian Platform (Odessa Shelf, Ukraine), crossing the Strandja, the Balkanide, the Moesian Platform and the Nord Dobrogea Orogen (compiled after *Finetti et al.* [1988], *Doglioni et al.* [1996], *Georgiev et al.* [2001], *Afanasenkov et al.* [2007], and the results of the present study). Note the crooked-line geometry of this transect crossing twice the same thrusted system of the Balkanides/Pontides. In order to visualize the same vergence of the thrusted Pontides system, the cross-section in the southern part is mirrored (dotted line). Note the concentration of thrust faults in zones with thinned continental crust (Moho depths after *Garkalenko* [1970] and *Grad and Tiira* [2009]) and high Tertiary thicknesses. Both panels display the same section, Figure 9 (top) has no vertical exaggeration, and Figure 9 (bottom) has 10X vertical exaggeration. For location see Figure 1.

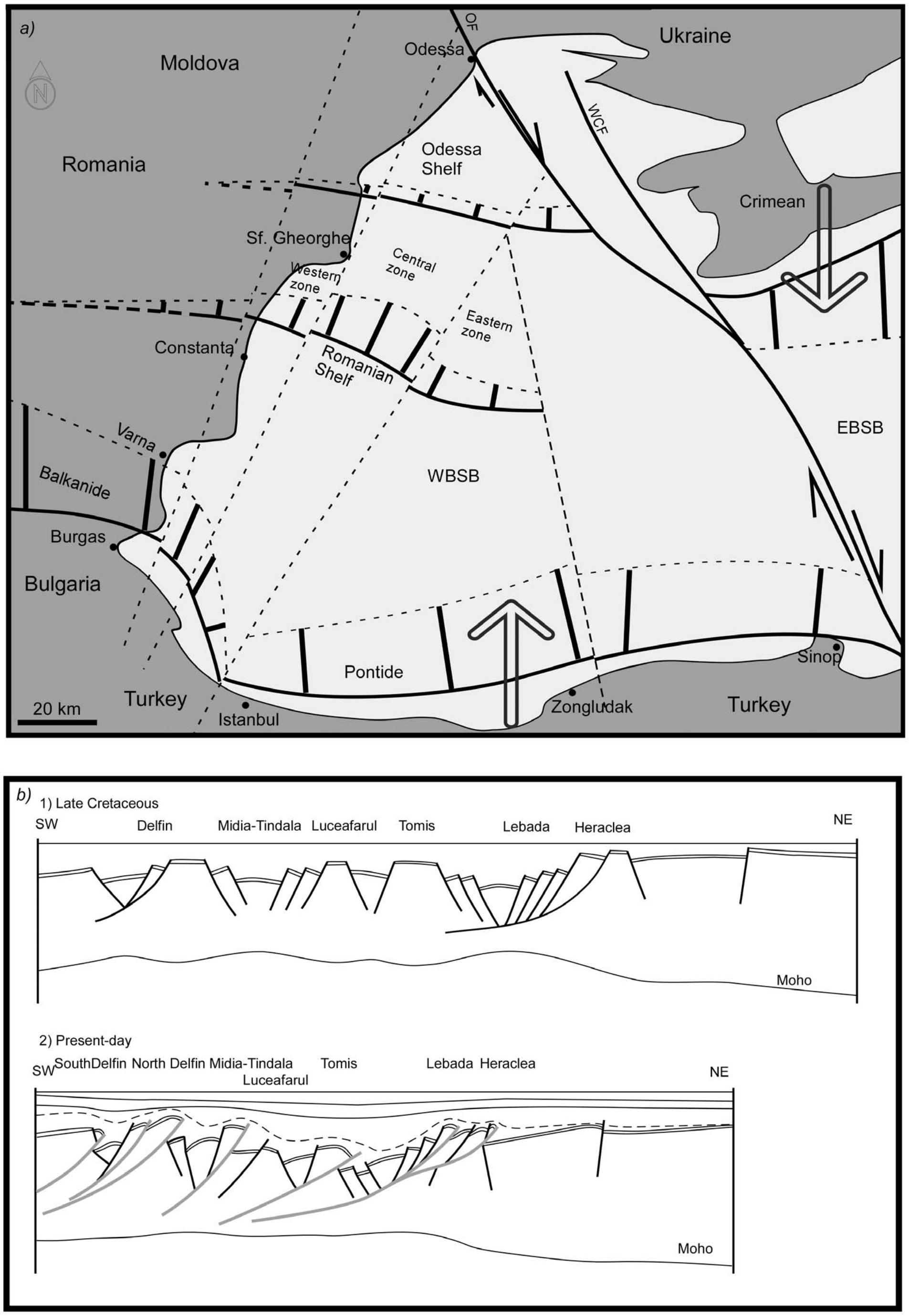



**Figure 11.** (a) Sketch of the kinematics of the late Eocene – middle Miocene inversion of the Black Sea. The western and eastern basins are decoupled in terms of deformation by the Odessa – West Crimean fault system. This system accommodates a change in the vergence of the thrust system, which is N-directed in the Western Black Sea and S-directed in the Eastern Black Sea. The vergence of the thrust system accommodating the shortening recorded between the Pontides and the Western Black Sea is taken from *Robinson and Kerusov* [1997]. The black thick continuous lines are estimated values for the amount of shortening (compiled after *Finetti et al.* [1988], *Robinson et al.* [1996], *Okay and Sahinturk* [1997], *Sinclair et al.* [1997], *Yilmaz et al.* [1997], *Georgiev et al.* [2001], *Menlikli et al.* [2009], *Khriachtchevskaia et al.* [2010], *Nikishin et al.* [2010], and results of this study). OF, Odessa Fault; WCF, West Crimen Fault. (b) Kinematic sketch illustrating the mechanism of inversion. Note the presence of graben – horst geometries, subsequently inverted by a thrust system overlain by syn-to post-kinematic deposits. The erosion of uplifted anticline culminations is not illustrated.

400 m-2 km individual shortening, cumulating a total in the order of 4 km (eastern zone in Figure 11).

[38] These variations in the amount of total shortening values recorded along the strike of the thrust system observed on the Romanian offshore and onshore can be correlated with similar, but opposite variations in the amounts of shortening calculated along the Balkanides-Western Pontides margin (Figure 11). The Balkanides offset is decreasing gradually from a shortening value of ~30 km in the area of the East Balkan unit to ~20 km in their coastal area and ~5–17 km in the offshore transition zone toward the Pontides (Figure 11). The vergence of the major structure accommodating the contraction observed between the Pontides and the Western Black Sea is not yet fully constrained by the association of large number of thrust faults with both N-ward and S-ward vergences. However, the total shortening observed increases in this sector to more than 25 km (Figure 11), no matter which interpretation is chosen (Figure 9). The northern extremity of the Black Sea records a total shortening of up to 3 km on the Odessa Shelf, gradually decreasing WNW-wards (Figure 11). This simple estimation of deformation amounts across the Western Black Sea and its immediate onshore demonstrate that the total shortening recorded is roughly constant along the strike of the thrust system in the order of ~30 km, lateral variation in thrusting being accommodated across the strike between the Romanian offshore-onshore and the Balkanide-Pontide system.

[39] Further to the east, the N-ward vergent thrusts observed on the Odessa Shelf are separated from the S-vergent Crimean-Caucasus Orogen by the Odessa-West Crimean fault system. The general characteristic of this system is the one of a wide transpressional area made up by small offset, high-angle reverse faults, along which occurred the dextral offset of the Odessa Shelf units relatively to the Crimea – East Black Sea Basin area. The Odessa-West Crimean fault system limit a horse-tail splay, which is located in the continuation of the western margin of the Mid- Black Sea High (Figure 1) [*Tugolesov et al.*, 1985; *Finetti et al.*, 1988; *Okay et al.*, 1994]. East of the West Crimean Fault, the Crimean – Great Caucasus system (Figures 10b and 10c) is part of the large scale S-ward thrusting observed at the northern margin of the Eastern Black Sea, with total shortening values in the order of 30 km during the late Eocene-Pliocene [e.g., *Shillington et al.*, 2008; *Nikishin et al.*, 2010, and references therein]. This shows a contrasting change in thrusting geometries between the Western and Eastern Black Sea domains: an N-ward vergent system with an indentor located S-wards (Pontides-Balkanides) is replaced eastward by an S-ward vergent system with an indentor located N-wards (Crimea-Greater Caucasus).

### 4.3. Shortening and Syn- to Post-kinematic Sedimentation Along the Romanian Offshore

[40] The late Eocene – middle Miocene kinematics of shortening is directly controlled by the pre-existence of Cretaceous grabens along the strike of the Romanian offshore. The concentration of extensional structures north of the Pontide-Balkanide indentor localized the thrusting in areas where the shortening direction was perpendicular to the strike of pre-existing grabens. The normal faults dipping southwards were inverted and the ones dipping northward were truncated (Figure 11a). In this respect, pre-existing grabens are weakness zones favoring localization of thrusting. In places where the direction of grabens was highly oblique to the direction of shortening, zero to minor inversion is recorded. For instance the large NE-SW oriented normal fault with offset in the order of 2 km observed offshore Varna (Figures 1 and 9) indicates little to no inversion [*Tari et al.*, 2009].

[41] Along the Romanian offshore, deformation truncated and uplifted in the hanging wall of thrusts the southern margins of all major Cretaceous grabens (Delfin, Midia-Tîndală, Tomis-Lotus, Luceafărul, Lebăda, Figure 11a). The uplift led to the partial to total exhumation and erosion that removed large portions of the syn-extensional basin fill. Thrusting was largely accompanied by subsidence, shortening structures being covered by Oligocene - middle Miocene syn- to post-kinematic sediments, particularly thick in the northern areas of the Histria Depression (e.g., Figure 3). The subsidence decreases WNW-wards to the onshore together with the above discussed decrease in thrusting offsets (Figure 11b). These sediments overlie both the footwall and the hanging wall of thrusts, the thrust-sheet top basins displaying an overall synclinal geometry (Figure 11a), which is common also for the foredeeps of the Balkanides and the Pontides (Figure 9). This geometry is in apparent contradiction with the theoretical wedge-shape models of foreland sedimentation [e.g., *Beaumont*, 1981]. However, this type of syn- to post-tectonic sedimentation covering the frontal part of thrust belts is rather common in Mediterranean-type of orogens (such as in the Apennines or the Carpathians), usually controlled by deep-scale processes or transmission of far-field-stresses localizing enhanced subsidence [e.g., *Bertotti et al.*, 2006; *Leever et al.*, 2006; *Picotti and Pazzaglia*, 2008]. In the Black Sea, the concentration of subsidence in areas previously affected by enhanced stretching suggests a causal relationship. It is likely that areas affected by extension have reset the thermo-tectonic age of the underlying lithosphere, its local weakening leading to crustal scale lithospheric folding (in the sense of *Cloetingh et al.* [1999]) at the same time with ongoing thrusting. Folding can be a result of far-field transmission of stresses from the Pontides and Balkanides indentation, similar to what has been proposed for the neighboring Carpathian Mountains [e.g., *Matenco et al.*, 2007]. Hence, the absence of numerical modeling renders our hypothesis speculative.

## 5. Conclusions

[42] The data presented in this study demonstrate that the structural highs observed by exploration studies offshore Romania form a coherent thick-skinned thrust system with N-ward vergence. Syn-kinematic patterns demonstrate that previously observed “depressions” (e.g., Histria) with unclear genesis cannot be related to processes such as extensional-related thermal subsidence as previously inferred [e.g., *Dinu et al.*, 2005]. These areas are foreland and thrust-sheet top basins buried by the subsidence taking place coeval with thrusting. Thrusting is focused in and inverts an existing geometry made up by successive grabens that are inherited from the Cretaceous extensional evolution. This extension started during Barremian and accelerated during Aptian-Albian

times, as demonstrated by the presence of Barremian synkinematic sedimentation, in agreement with previous observations made elsewhere in the West Black Sea [e.g., *Okay et al.*, 1994; *Robinson et al.*, 1996; *Robinson and Kerusov*, 1997; *Spadini et al.*, 1997; *Hippolyte et al.*, 2010]. The extension continued throughout the remaining Cretaceous times with fairly similar effects in terms of structural WNW-ESE oriented trends. One particular feature is the Peceneaga-Camena Fault which functioned as a transcurrent system accommodating differential amounts of stretching and creating rapidly uplifting of basement blocks at restraining strike-slip bends.

[43] This study demonstrates that the shortening starting with the late Eocene affected all areas of the Western Black Sea, deformation being coherently correlated across its western margin. The geometry of thrusts (Figure 9) and the decreasing offsets northward demonstrate that the cause of this generalized inversion is linked with transmission of stresses during what is known as the final part of the Cretaceous – Eocene collision recorded in the Pontide-Balkanide system, related to the N-ward subduction and closure of the Tethys ocean [e.g., *Şengör*, 1987]. Syn-tectonic sedimentation in the Western Black Sea Basin demonstrates that this process was continuous in the back-arc and took place through the onset of gradual shortening migrating northward. Although the total amount of shortening is roughly constant in an E-W direction, thrusts have a highly variable offset along their strike, deformation amounts being transferred between individual structures located at distance across the strike of the system (Figure 11). The geometry of the N-ward vergent thrust system of the Western Black Sea is incompatible with the S-ward vergent thrusting driven by the Crimean-Caucasus system of the Eastern Black Sea. Therefore, the Odessa-West Crimea fault system must have accommodated substantial amounts of dextral deformation, variable along their strike (maximum near the margins and minimum in the center of the basin, Figure 11b).

[44] The continuation of shortening pointed out in our study, show that the process of Tethyan collision did not stop during Eocene times as previously suggested, but continued in the back-arc much later until middle Miocene times. Whether or not this ongoing thrusting can be connected in all areas of the Western Black Sea with deeper-seated, lower crustal thrust faults (Figure 9) is a matter of further research. However, this is interesting in the light of the Miocene exhumation of Pontides correlated with current models of Anatolian Plateau being mantle supported during coeval shortening [e.g., *Şengör et al.*, 2003].

[45] The Black Sea example demonstrates the mechanism of transition from an extensional back-arc basin to a contractional one during the evolution of the Tethyan subduction and collision. Such back-arc transitions can be furthermore correlated with changing back-arc stress regimes during the oceanic subduction observed elsewhere, such as the variability along the Andean subduction zone caused by fine variation in subduction parameters [e.g., *Sobolev et al.*, 2007; *Gerbault et al.*, 2009]. In the case of the Black Sea, while the extension was probably associated with processes such as slab retreat, the inversion was possibly associated with the onset of buoyant accumulation of lower plate crustal material at the subduction zone during collision. The amount of back-arc deformation recorded by the later was rather reduced, with a significant exhumation of various crustal blocks due to involvement of thick-skinned structures.

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